UCRL-JRNL-207218

Quantifying the effects of CO2-fertilized vegetation on future global climate and carbon dynamics

S. L. Thompson, B. Govindasamy, A. Mirin, K. Caldeira, C. Delire, J. Milovich, M. Wickett, D. Erickson

October 14, 2004

Geophysical Research Letters

Disclaimer

Quantifying the effects of CO_2-fertilized vegetation on future global climate and carbon dynamics

Starley L. Thompson[1], Bala Govindasamy[1], Art Mirin[1], Ken Caldeira[1], Christine Delire[2], Jose Milovich[1], Mike Wickett[1], David Erickson[3]

[1]Lawrence Livermore National Laboratory
Livermore, California, USA

[2] University of Wisconsin-Madison
Present address: Institut des Sciences de l'Evolution, Université Montpellier II, France

[3]Oak Ridge National Laboratory
Oak Ridge, Tennessee, USA

Abstract

Climate and the global carbon cycle are a tightly coupled system where changes in climate affect exchange of atmospheric CO_2 with the land biosphere and the ocean, and vice-versa. In particular, the response of the land biosphere to the ongoing increase in atmospheric CO_2 is not well understood. To evaluate the approximate upper and lower limits of land carbon uptake, we perform simulations using a comprehensive climate-carbon model. In one case the land biosphere is vigorously fertilized by added CO_2 and sequesters carbon throughout the 21st century. In a second case, CO_2 fertilization saturates in year 2000; here the land becomes an additional source of CO_2 by 2050. The predicted atmospheric CO_2 concentration at year 2100 differs by 40% between the two cases. We show that current uncertainties preclude determination of whether the land biosphere will amplify or damp atmospheric CO_2 increases by the end of the century.

Introduction

The magnitudes of feedbacks within the climate-carbon system are poorly constrained. Higher CO_2 concentrations increase photosynthesis and promote water-use and nitrogen-use efficiency of plants, ultimately increasing plant growth (*Houghton*, *et al.,* 2001). Biomass and soil carbon, and thus terrestrial carbon uptake, may be expected to increase with higher atmospheric CO_2 levels. However, the effects of photosynthetic CO_2 "fertilization" will saturate at sufficiently high CO_2 levels (*Farquhar, et al.,* 1980; *Houghton, et al.*, 2001), and higher global temperatures may increase the loss of soil carbon to the atmosphere (*Giardina and Ryan*, 2004; *Houghton, et al.*, 2001; *Lloyd and Taylor*, 1994).

Results from two recent modeling studies, referred to here as Hadley (*Cox, et al.*, 2000) and IPSL (*Friedlingstein, et al.*, 2001), led to different conclusions regarding the role of the land biosphere in future global change. In the Hadley simulation, the land biosphere becomes a net source of CO_2 to the atmosphere by year 2050, whereas in the IPSL simulation, it remains a net sink throughout the 21st century. We show that we can produce a change of sign in biospheric response by changing only one assumption in a fully coupled three-dimensional model: whether CO_2-fertilization rapidly saturates in terrestrial ecosystems. Other factors, such as climate change, land-use change, fire suppression, and N-fertilization could also potentially affect the sign of terrestrial carbon uptake.

Higher atmospheric CO_2 concentration stimulates leaf-photosynthesis and favors more efficient use of available water (*Owensby, et al.*, 1999). Models incorporating this dynamic without nutrient constraints to growth tend to be more sensitive to CO_2 fertilization (*Cramer, et al.*, 2001; *Kicklighter, et al.*, 1999). However, in real ecosystems, limited availability of nutrients or long-term acclimation responses may diminish the sensitivity to added CO_2 (*Hungate, et al.*, 2003; *Nadelhoffer, et al.*, 1999; *Shaw, et al*., 2002; *Schimel*, 1998).

Coupled Climate/Carbon Model

To investigate the dynamics of the land biosphere in the coupled climatic system, we coupled the ocean-atmosphere climate model PCTM (*Meehl, et al*., 2004; *Washington, et al*., 2000) to the IBIS2 terrestrial biosphere model (*Foley, et al*., 1996; *Kucharik, et al.*, 2000) and a modified version of the OCMIP ocean biogeochemistry model protocol (*Najjar and Orr*, 1999). The horizontal resolutions of the atmosphere and ocean circulation model grids are approximately 280 km and 70 km, respectively. IBIS2 simulates canopy physiology, plant phenology, vegetation dynamics and carbon cycling. The ocean biogeochemistry model predicts air-sea CO_2 fluxes, biogenic export of organic matter and calcium carbonate, and distributions of dissolved inorganic carbon, phosphate, oxygen, alkalinity, and dissolved organic matter. We replaced the OCMIP export formulation with a formulation based on that of *Maier-Reimer* (1993).

In the LLNL PCTM-IBIS2-OCMIP system, precipitation biases typical of climate models can cause vegetation errors that, in turn, amplify precipitation biases. This erroneous feedback results in unacceptable vegetation in some areas, particularly parts of the Amazon. To remedy this, at every surface grid cell and time step the precipitation field is multiplied by a correction field that is a function of position, but otherwise static and identical across all runs. We further scale the precipitation field so as to maintain the global conservation of water. This correction procedure spatially redistributes the model's present-day annual mean precipitation towards observed means.

Experiments

We integrate the fully coupled model to quasi-equilibrium to form a year 1870 pre-industrial initial condition that is used as the starting point for three model cases. The "control" case has no CO_2 emissions and thus no change in radiative forcing for the period 1870-2100. Model drift for the control case over years 1900 to 2100 is -0.35 K in mean surface temperature and +3.14 ppmv in atmospheric CO_2. Both are minimal residuals from an imbalance in the initial state and are not subtracted from the other simulations in our analysis.

The "fertilization" case has CO_2 emissions specified at historical levels (*Marland, et al.*, 2000) for 1870-2000 and that follow the IPCC scenario (*Houghton, et al.*, 2001) SRES A2 from

2000-2100. Non-CO_2 greenhouse gas concentrations are specified at historical levels for 1870-2000 and SRES A2 level from 2000-2100. Land use emissions are reconstructions (*Houghton*, 2003) for the historical period and from the SRES A2 scenario thereafter. In this scenario, total emissions reach 29 Gigatons carbon (GtC) per year in year 2100 from present day values of 8 GtC per year.

The "saturation" case is identical to the fertilization case except the CO_2 fertilization is assumed to saturate at the year 2000 concentration, accomplished by forcing the terrestrial biosphere model with a constant CO_2 concentration of 366 ppmv after year 2000. Because the terrestrial biosphere model that we use is highly responsive to CO_2 fertilization (*Cramer, et al*., 2001), the fertilization case approximates an upper limit to the land uptake of carbon assuming unlimited nitrogen/nutrient availability. Capping all fertilization at its year 2000 value in the saturation case will approximate a strongly nitrogen/nutrient limited system.

Results

Figure 1a shows that assumptions regarding CO_2-fertilization of the land biosphere greatly affect the atmospheric concentration of CO_2. Year 2100 atmospheric CO_2 concentrations are 336 ppmv higher in the saturation case than in the fertilization case. In the SRES A2 scenario, 1790 GtC is emitted to the atmosphere over the 21st century; atmospheric CO_2 content increases by 776 GtC (366 ppmv) and 1489 GtC (702 ppmv) in our fertilization and saturation cases, respectively.

The global climate-carbon feedback factor is a useful system metric defined as the ratio of CO_2 change when climate is changing, to the CO_2 change when climate is constant (*Friedlingstein, et al*., 2003). We perform an additional constant-climate simulation with full emissions to obtain a feedback factor of 1.13 for our fertilization case. The feedback factors for similar fertilization simulations are 1.19 for IPSL (*Friedlingstein, et al.*, 2001) and 1.68 for Hadley (*Cox, et al.*, 2000). Therefore, our model shows the weakest positive feedback between climate and the carbon cycle of the current published results for fertilization cases. Note, however, that our feedback factor increases to 2.05 in our saturation case. This is an indication of the uncertainty in quantifying the climate-carbon cycle feedback, arising from a single model assumption.

The temperature difference at year 2100 between the saturation and fertilization cases is only 0.7 K (Fig. 1b), but it should be noted that the climatic system has large thermal inertia due to the large heat capacity of the oceans. If the simulations were run to equilibrium with the year 2100 CO_2 values, the temperature difference would be approximately 1.1 K (estimated from the model's equilibrium climate sensitivity of 2.1 K per doubling of CO_2).

Results (Fig. 2a) show that assumptions regarding the saturation of CO_2-fertilization can affect the sign of the CO_2 flux between the atmosphere and land by century's end. Direct CO_2 effects are expected to lead to increased terrestrial carbon uptake, but temperature effects can lead to increased heterotrophic respiration and loss of soil carbon (*Cox, et al.*, 2000; *Cramer, et al.*, 2001; *Friedlingstein, et al.*, 2001; *Joos, et al.*, 2001), at least until a possible acclimation of soil microbiology to the higher temperatures (*Kirschbaum*, 2000; *Tjoelker, et al.*, 2001). In the saturation case, by year 2100 the land-biosphere has become a net source of CO_2 to the atmosphere, as temperature effects dominate CO_2-fertilization effects. In the fertilization case, CO_2-fertilization effects dominate temperature effects, resulting in continued net biospheric growth.

In contrast to Hadley (*Cox, et al*., 2000), but in agreement with IPSL (*Friedlingstein, et al.*, 2001), our land carbon cycle does not become a net source of carbon to the atmosphere in the fertilization case. A loss of vegetation biomass does not occur in either of our simulations (but soil carbon does decline by year 2100 in our saturation case).

The model's historical ocean carbon uptake of 77 GtC is an underestimate when compared against the recent observational estimates of 118 GtC for the period from 1800 to 1994 (*Sabine et al., 2004*). Between year 2000 and year 2100, ocean/atmosphere carbon fluxes show significant differences between the fertilization and saturation cases (Fig. 2b). Ocean carbon storage increases by 269 and 357 GtC in the two cases (Fig 2c). Ocean uptake is greater in the saturation case because increased atmospheric CO_2 concentrations drive an increased flux of CO_2 from the atmosphere to the ocean that is larger than any counteracting temperature and biogeochemical effects (*Sarmiento, et al*., 1998). The increase in ocean carbon storage at year 2100 in both the cases is less than 15 % of the amount that could be stored if the ocean were in equilibrium with the respective year 2100 atmospheric CO_2 concentrations (*Kheshgi*, 2004).

Cumulative carbon emissions since 1870 reach 2200 GtC by year 2100 (Fig. 2c). In the fertilization case, the land biosphere and the oceans sequester 919 GtC (42%) and 346 GtC (15.5%) of the total emissions, respectively. In the saturation case, the corresponding amounts are 104 GtC (5%) and 435 GtC (19.5%). Thus, in our model the variation in land uptake of carbon due to the degree of CO_2 fertilization varies from 5% to 42% of the total carbon emitted. The carbon remaining in the atmosphere is 935 GtC (42.5%) and 1661 GtC (75.5%) for the fertilization and saturation cases, respectively.

In the fertilization case, the total land ecosystem nitrogen increases by 20 Gt between year 2000 and 2100. This is much larger than estimates (*Hungate, et al*., 2003), indicating only 6

Gt of additional nitrogen could accumulate in the terrestrial biosphere by 2100. In contrast, in the saturation case terrestrial nitrogen declines by 8 Gt during the same period.

The geography of simulated carbon uptake in the fertilization case over the period 1870-2100 (Fig. 3a) shows that anthropogenic carbon is stored on land primarily in areas of high vegetation productivity (Amazonia, central Africa, south and southeast Asia) or relatively slow microbial decomposition (boreal forests). Currents and circulation make storage somewhat more uniform for the ocean, but it is higher in the North Atlantic and Mid-Southern Oceans, which reflects proximity to regions of net CO_2 uptake (*Caldeira and Duffy,* 2000; *Wickett, et al.*, 2003). In the saturation case (Fig. 3b), the oceanic storage increases slightly and carbon is released over many parts of the land area.

Discussion

Even without nutrient limitations, the enhanced physiological effects of CO_2 on plant growth will saturate at high CO_2 concentration (*Cao and Woodward*, 1998; *Farquhar, et al.*, 1980; *Houghton, et al.*, 2001). However, respiration rates would be expected to continue to increase with further CO_2-induced warming; thus, at some level of atmospheric CO_2, the land would be expected to become a net CO_2 source. Questions remain when this would occur, and whether it would occur before fossil fuel resources are fully exploited. The climate model used here has an equilibrium climate sensitivity that is at the lower end of the range of the general model population (*Houghton, et al.*, 2001). A more sensitive climate model would increase the amount of warming, increasing heterotrophic respiratory carbon fluxes from soils even more. Hence, high climate sensitivity is more likely to amplify carbon losses from the land biosphere; a low climate-sensitivity is more likely to allow the land biosphere to damp the climate effects of CO_2 emissions, with carbon uptake by the biosphere dominated by CO_2 fertilization.

For the range of CO_2 fertilization considered in this study, the sign of the net flux of carbon between the atmosphere and land biosphere changes by the end of this century. It is possible that CO_2-fertilization is weaker (or stronger) than in our simulations, and thus the shift of the land biosphere from a net atmospheric CO_2 sink to net atmospheric CO_2 source could occur earlier (or later) than simulated here.

The choice of year 2000 for the saturation level is artificial; it is driven by the need to have the model reproduce the past evolution of atmospheric CO_2. The processes responsible for the observed land uptake are uncertain. A standard hypothesis uses the CO_2-fertilization effect, represented by our fertilization case. The observed uptake could be due to other processes such as change in land use, N-fertilization and fire suppression (*Pacala et al.*, 2001; *Schimel et al.*, 2001). This is represented by our saturation case where the uptake is, for practical reason only, driven by the atmospheric CO_2 for the historical period. The behavior of CO_2-fertilization and other processes in the future is uncertain

The results show that the amount of anthropogenic CO_2 in the atmosphere at the end of the century will probably be sensitive to carbon-cycle processes about which we are uncertain at present. Right now, the uncertain response of the land biosphere to increased CO_2 and climate change prevents us from knowing if terrestrial carbon-cycle feedbacks will damp or amplify global warming.

Acknowledgements

This work was performed under the auspices of the U.S. Department of Energy by the

University of California Lawrence Livermore National Laboratory under contract No. W-7405-Eng-48. Thanks to W.M. Washington and his team for use of the PCTM model, and to J. Foley for IBIS2. This is LLNL report UCRL-JRNL-207218.

References

Caldeira, K. and P.B. Duffy (2000), The role of the Southern Ocean in uptake and storage of anthropogenic carbon dioxide, *Science*, 287, 620-622.

Cao, M.K. and F.I. Woodward (1998), Dynamic responses of terrestrial ecosystem carbon cycling to global climate change, *Nature*, 393, 249-252.

Cox, P.M., R.A. Betts, C.D. Jones, S.A. Spall and I.J. Totterdell (2000), Acceleration of global warming due to carbon-cycle feedbacks in a coupled climate model, *Nature*, 408, 184-187.

Cramer, W., *et al.* (2001), Global response of terrestrial ecosystem structure and function to CO_2 and climate change: results from six dynamic global vegetation models, *Global Change Biology*, 7, 357-373.

Farquhar, G.D., S.V. Caemmerer and J.A. Berry (1980), A biochemical-model of photosynthetic CO_2 assimilation in leaves of C-3 species, *Planta*, 149, 78-90.

Foley, J.A., *et al.* (1996), An integrated biosphere model of land surface processes, terrestrial carbon balance, and vegetation dynamics, *Global Biogeochem. Cycles*, 10, 603-628.

Friedlingstein, P, *et al*. (2001), Positive feedback between future climate change and the carbon cycle, *Geophys. Res. Lett.*, 28, 1543-1546.

Friedlingstein, P., J.-L. Dufresne, P.M. Cox and P. Rayner (2003), How positive is the feedback between climate change and the carbon cycle? *Tellus*, 55B, 692-700.

Giardina, C.P. and M.G. Ryan (2000), Evidence that decomposition rates of organic carbon in mineral soil do not vary with temperature, *Nature*, 404, 858-861.

Houghton, R.A (2003), Revised estimates of the annual net flux of carbon to the atmosphere from changes in land use and land management 1850-2000, *Tellus*, 55B, 378-390.

Houghton, J.T. *et al.,* (Eds.) (2001), *Climate Change 2001, The Scientific Basis*, Cambridge Univ. Press, Cambridge.

Hungate, B.A., J.S. Dukes, M.R. Shaw, Y. Luo and C.B. Field (2003), Nitrogen and climate change, *Science*, 302, 1512-1513.

Joos, F., *et al.* (2001), Global warming feedbacks on terrestrial carbon uptake under the Intergovernmental Panel on Climate Change (IPCC) emission scenarios, *Global Biogeochemical Cycles*, 15, 891-907.

Kheshgi, H. S., Ocean carbon sink duration under stabilization of atmospheric CO_2 – A 1000 year time scale, *Geophys. Res. Lett*. (in press).

Kicklighter, D.W., *et al.* (1999), A first-order analysis of the potential role of CO_2 fertilization to affect the global carbon budget: a comparison of four terrestrial biosphere models, *Tellus*, 51B, 343-366.

Kirschbaum, M.U.F. (2000), Will changes in soil organic carbon act as a positive or negative feedback on global warming? *Biogeochemistry*, 48, 21-51.

Kucharik, C.J., *et al* (2000)*,* Testing the performance of a Dynamic Global Ecosystem Model: Water balance, carbon balance, and vegetation structure, *Global Biogeochem. Cycles* 14 (3), 795-825.

Lloyd, J. and J.A. Taylor (1994), On the temperature-dependence of soil respiration, *Functional Ecology*, 8, 315-323.

Maier-Reimer, E. (1993), Geochemical cycles in an ocean general-circulation model – preindustrial tracer distributions, *Global Biogeochemical Cycles,* 7(3), 645-677.

Marland, G., T. Boden and R. Andres (2000), Global, regional, and national annual CO_2 emissions from fossil-fuel burning, cement production and gas flaring: 1751-1999, *CDIAC NDP-030*, Carbon Dioxide Information Analysis Center, Oak Ridge National Laboratory, Oak Ridge, Tennessee.

Meehl, G.A., W.M. Washington, J.M. Arblaster. and A.X. Hu (2004), Factors affecting climate sensitivity in global coupled models, *J. Climate* 17, 1584-1596.

Nadelhoffer, K.J., *et al.* (1999), Nitrogen deposition makes a minor contribution to carbon sequestration in temperate forests, *Nature*, 398, 145-148.

Najjar, R.G. and J.C. Orr (1999), Biotic How-To, Revision 1.7, Ocean Carbon-cycle Model Intercomparison Project (OCMIP), http://www.ipsl.jussieu.fr/OCMIP/phase2/simulations/Biotic/HOWTO-Biotic.html

Owensby, C.E., J.M. Ham, A.K. Knapp and L.M. Auen (1999), Biomass production and species composition change in a tallgrass prairie ecosystem after long-term exposure to elevated atmospheric CO_2*, Global Change Biology*, 5, 497-506.

Pacala, S.W., *et al*. (2001), Consistent land- and atmosphere-based U.S. carbon sink estimates, *Science*, 292, 2316-2320.

Sabine, C.L., *et al*. (2004), The oceanic sink for anthropogenic CO_2, *Science*, 305, 367-371.

Sarmiento, J.L, M.C. Hughes, R.J. Stouffer and S. Manabe (1998), Simulated response of the ocean carbon cycle to anthropogenic climate warming, *Nature*, 393, 245-249.

Schimel, D.S (1998), Climate change - The carbon equation, *Nature*, 393, 208-209.

Schimel, D.S., *et al*. (2001), Recent patterns and mechanisms of carbon exchange by terrestrial ecosystems, *Nature*, 414, 169-172.

Shaw, M.R., *et al*. (2002), Grassland responses to global environmental changes suppressed by elevated CO_2. *Science*, 298, 1987-1990.

Tjoelker, M.G., J. Oleksyn and P.B. Reich (2001), Modelling respiration of vegetation: evidence for a general temperature-dependent Q(10), *Global Change Biology*, 7, 223-230.

Wickett, M.E., K. Caldeira and P.B. Duffy (2003), Effect of horizontal grid resolution on simulations of oceanic CFC-11 uptake and direct injection of anthropogenic CO_2, *J. Geophys. Res. (Oceans)*, 108, art. no. 3189.

Washington, W.M., *et al.* (2000), Parallel climate model (PCM) control and transient simulations, *Climate Dynamics*, 16, 755-774.

Figure Captions

Figure 1. Simulated atmospheric CO_2 and global mean surface temperature from 1870 to 2100. (a) CO_2 for the control (black), fertilization (green), and saturation (red) cases. Black dots are observed CO_2 concentrations. If CO_2 fertilization saturates early, the land-biosphere becomes a net source of CO_2 to the atmosphere, adding to anthropogenic CO_2 emissions. (b) Temperature for the same cases as (a).

Figure 2. Global carbon fluxes and carbon inventory change from 1870 to 2100. (a) Flux of carbon from land to atmosphere. Control (black), fertilization (green), and saturation (red) cases. In the saturation case the land becomes a net source of carbon by year 2050. (b) The same as (a) but for carbon flux from ocean to atmosphere. (c) Global carbon change from the 1870 pre-industrial starting point. Total earth system (black), land (solid), and ocean (dashed). Fertilization case (green), and saturation case (red).

Figure 3. The simulated geography of carbon stored in the earth system over the period from 1870 to 2100 in the a) fertilization and b) saturation cases (column integrated carbon in kg C / m^2). Anthropogenic carbon is stored primarily in areas of high vegetation productivity and/or cooler climates over land. Owing to currents, storage is somewhat more uniform for the oceans but higher in the North Atlantic and Mid-Southern oceans, which reflects proximity to regions of net CO_2 uptake. In the saturation case, the oceanic storage increases slightly and carbon is released over many parts of the land area.

Figure 1 Simulated atmospheric CO_2 and global mean surface temperature from 1870 to 2100. (a) CO_2 for the control (black), fertilization (green), and saturation (red) cases. Black dots are observed CO_2 concentrations. If CO_2 fertilization saturates early, the land-biosphere becomes a net source of CO_2 to the atmosphere, adding to anthropogenic CO_2 emissions. (b) Temperature for the same cases as (a).

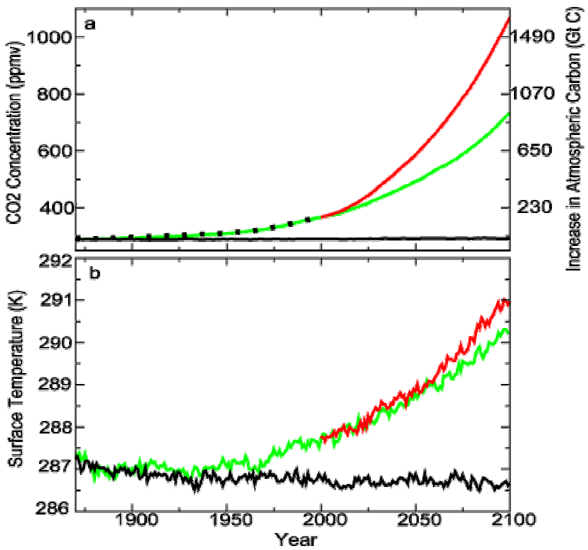

Figure 2 Global carbon fluxes and carbon inventory change from 1870 to 2100. (a) Flux of carbon from land to atmosphere. Control (black), fertilization (green), and saturation (red) cases. In the saturation case the land becomes a net source of carbon by year 2050. (b) The same as (a) but for carbon flux from ocean to atmosphere. (c) Global carbon change from the 1870 pre-industrial starting point. Total earth system (black), land (solid), and ocean (dashed). Fertilization case (green), and saturation case (red).

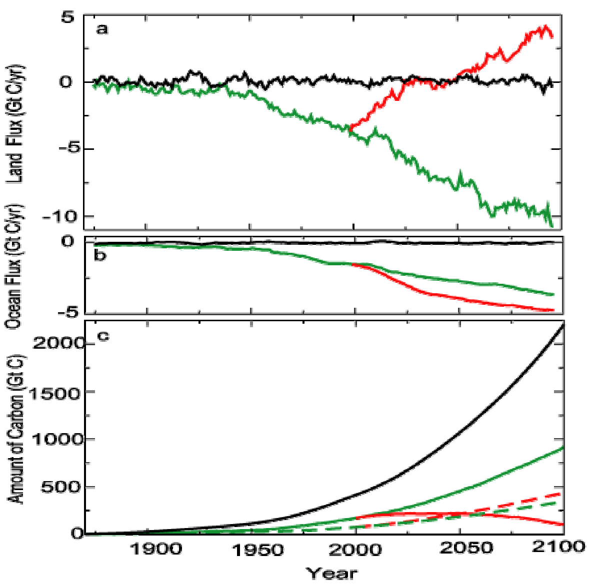

Figure 3 The simulated geography of carbon stored in the earth system over the period from 1870 to 2100 in the a) fertilization and b) saturation cases (column integrated carbon in kg C / m^2). Anthropogenic carbon is stored primarily in areas of high vegetation productivity and/or cooler climates over land. Owing to currents, storage is somewhat more uniform for the oceans but higher in the North Atlantic and Mid-Southern oceans, which reflects proximity to regions of net CO_2 uptake. In the saturation case, the oceanic storage increases slightly and carbon is released over many parts of the land area.